Application of Rouse’s Sediment Concentration Profile to Aeolian Transport: Is the suspension system for sand transport in air the same as that in water?

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ABSTRACT

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In this study, the applicability of Rouse’s suspended sediment concentration profile for fluvial transport to aeolian transport was shown and a universal system of suspended sand transport in water and air was investigated, in order to obtain knowledge of suspended sand transport in air. Comparison between Rouse’s vertical concentration profile and the profile obtained from wind tunnel experimental results indicated that Rouse’s profile underestimated the experimental results when the mode of sand particle motion was classified into saltation; however, it agreed with the experimental results when the mode of sand particle motion was classified into suspension under strong wind conditions. The mode of sand particle motion was universally classified on the basis of a dimensionless Rouse number for both fluvial and aeolian transports. Results of this study suggested that Rouse’s profile for fluvial transport is possibly applicable to aeolian transport when the Rouse number is roughly greater than 1 under strong wind conditions.

ADDITIONAL INDEX WORDS: dimensionless concentration profile, transport mode, density ratio, particle velocity, saltation layer

INTRODUCTION

Wind erosion is one of the serious problems in arid and semi-arid regions. In the vicinity of sandy beaches, aeolian sand causes problems such as sand deposition on the road and sand transportation into agricultural fields. Sand is transported by various modes that include saltation and suspension. The most significant difference between saltation and suspension systems is that suspension focuses on a sediment diffusion process while saltation focuses on sediment splash and rebound processes. Most existing studies on aeolian sand transport have focused on only the saltation system, and not the suspension system.

Many studies have been conducted on aeolian sand transport in saltation; in addition, from the results of wind tunnel experiments and field observations, several formulae on the saltation flux have been proposed (e.g., Bagnold, 1941; Kawamura, 1964). However, few studies have been conducted on that in suspension (without distinction between saltation and suspension; Gillette and Chen, 1999; Pasini and Jenkins, 2005; Anderson and Walker, 2006), while several formulae for fluvial transport on the suspension flux have been proposed (e.g., Rouse, 1937; Itakura and Kishi, 1980) as well as the saltation flux (e.g., Meyer-Peter and Muller, 1948). This is because the main transport mode of sand particles seems to be saltation for aeolian sand, and it is both saltation and suspension for fluvial sand. Because of a theoretical difference between saltation and suspension fluxes, the theory of saltation is not applicable to suspension, and the aeolian suspension needs another formulation.

Thus far, several researchers have studied both fluvial and aeolian sand transports because these types of sands differ only in terms of their particle-to-fluid density ratios. There are several differences between fluvial and aeolian transports such that the dimensionless threshold, i.e., Shields parameter decreases with an increase in the density ratio, and transport length of sand increases for a small density ratio (i.e., fluvial transport) and decreases for a large density ratio (i.e., aeolian transport) (Iversen et al., 1987; Tsuchiya, 1991); however, there are several common characteristics between fluvial and aeolian transports, such as the logarithmic fluid profile law can be applied to both, the threshold Shields parameter has a minimal value when the dimensionless particle diameter $D_p$ is approximately equal to 10, and the transport mode is classified on the basis of the Rouse number, which depends on the settling velocity and fluid shear velocity (Van Rijn, 1993; Hu and Hui, 1996; Dade and Friend, 1998; and Shao, 2000). These facts indicate that the knowledge of fluvial transport in suspension might be applicable to aeolian transport.

In order to elucidate the differences between sand transport mechanisms in water and air and to investigate the aeolian suspended sand transport flux, the concentration profiles of aeolian sand transport flux calculated using the experimental data obtained by Ni et al. (2002) and Dong et al. (2006) was compared with Rouse’s (1937) concentration profile which was derived as a fluvial sand concentration profile from the diffusion equation.
Table 1: Summary of experimental conditions (Ni et al., 2002).

<table>
<thead>
<tr>
<th>Case</th>
<th>$d$</th>
<th>$U$</th>
<th>$u^*$</th>
<th>$z_0$</th>
<th>$H$</th>
<th>$N$</th>
</tr>
</thead>
<tbody>
<tr>
<td>17O</td>
<td>0.17</td>
<td>-</td>
<td>0.19 ($u_*$)</td>
<td>0.05</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>17A</td>
<td>0.17</td>
<td>8.5</td>
<td>0.60</td>
<td>0.77</td>
<td>0.19</td>
<td>4.65</td>
</tr>
<tr>
<td>17B</td>
<td>0.17</td>
<td>11.5</td>
<td>0.90</td>
<td>1.75</td>
<td>0.29</td>
<td>3.10</td>
</tr>
<tr>
<td>17C</td>
<td>0.17</td>
<td>13.5</td>
<td>1.20</td>
<td>2.37</td>
<td>0.33</td>
<td>2.33</td>
</tr>
<tr>
<td>17D</td>
<td>0.17</td>
<td>16.5</td>
<td>1.73</td>
<td>7.86</td>
<td>0.34</td>
<td>1.62</td>
</tr>
<tr>
<td>17E</td>
<td>0.17</td>
<td>22.5</td>
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<td>9.56</td>
<td>0.35</td>
<td>1.20</td>
</tr>
<tr>
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<td>-</td>
<td>0.30 ($u_*$)</td>
<td>0.05</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
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<td>0.48</td>
<td>0.39</td>
<td>0.31</td>
<td>10.30</td>
</tr>
<tr>
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<td>11.5</td>
<td>0.78</td>
<td>1.82</td>
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<td>6.37</td>
</tr>
<tr>
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<td>0.35</td>
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<td>3.27</td>
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<td>4.37</td>
</tr>
<tr>
<td>35D</td>
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<td>8.67</td>
<td>0.38</td>
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</tr>
<tr>
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<td>2.38</td>
<td>11.75</td>
<td>0.41</td>
<td>2.09</td>
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</table>

**ROUSE’S SEDIMENT CONCENTRATION PROFILE**

The vertical sand mass flux at a specific height is defined as the mass passing a unit cross-sectional area in unit time. This flux can be obtained from the following equation if the corresponding particle concentration $C(z)$ and mean particle velocity $u_p(z)$ are known.

$$q(z) = C(z) u_p(z)$$  \hspace{1cm} (1)

$u_p(z)$ in water is usually expressed in terms of the fluid velocity $u(z)$. In this study, Rouse’s (1937) equation was selected from several suspended sediment transport equations for fluvial sand transport, because it is the simplest equation. Rouse’s concentration profile was derived from the diffusion equation by considering the effect of the settling velocity of sand, and it was expressed as

$$\frac{C(z)}{C(a)} = \left[ \frac{h-z}{z} \right]^{-k}$$ \hspace{1cm} \frac{N = \frac{w_0}{\kappa u_*}}{ (2)}$$

where $C(z)$ is the concentration at height $z$ above the mean bed, $C(a)$ is the reference concentration at a height $z = a$ of the saltation layer, $h$ is the water depth, $N$ is the Rouse number, $w_0$ is the settling velocity, $\kappa$ is the von Karman constant ($=0.4$), and $u_*$ is the bed shear velocity. The value of $a$ on which the vertical profile depends significantly is generally given by 0.05$h$ (Vanoni, 1946), half the bed form height (Van Rijn, 1984), or 10$h$/$d$ ($d$ is the diameter of a sand particle; Shibayama and Rattanapitikon, 1993). The value of $w_0$ is determined using Rubey’s (1933) expression by

$$w_0 = \sqrt{s-1}gd \left\{ \frac{2 + 36\nu^2}{3} \right\}^{\sqrt{\frac{36\nu^2}{(s-1)gd^2}}} \right\}$$ \hspace{1cm} (4)

where $s$ is the relative density of sand ($=\sigma d \rho$, where $\sigma$ is the sand density), $g$ is the gravity acceleration, and $\nu$ is the kinematic viscosity of fluid.

**Figure 1.** Vertical profiles of wind velocity $u$, sand mass flux $q$, mean particle velocity $u_p$, and concentration $c$ for (a) sand A and (b) sand B (data obtained from Ni et al. (2002)).

Rouse’s concentration profile was verified by Vanoni’s (1946) experimental results for fluvial sand transport when the diameters of the sand particles ranged from 0.09 to 0.15 mm and the fluid velocities ranged from 0.74 to 2.00 m s$^{-1}$, i.e., the Rouse number ranged from 0.34 to 1.46.

**SAND CONCENTRATION PROFILE FOR AEOLIAN TRANSPORT**

In order to compare the vertical concentration profile for aeolian transport with Rouse’s profile, $C(z)$ should be calculated using $q(z)$ and $u_p(z)$, which are measured experimentally (equation (1)). Several studies have been conducted on the vertical distribution of aeolian sand flux by performing wind tunnel experiments (e.g., Ni et al., 2002; Kubota et al., 2006) and field experiments (Namikas, 2003). Here, $q(z)$ obtained from the wind tunnel experiments by Ni et al. (2002) was used because $u_p(z)$ was also obtained from experiments using the same wind tunnel by Dong (2006). These experiments were performed in a straight-line blowing wind tunnel at the Shapotou Desert Research Station (Chinese Academy of Sciences). The test section was 21 m long, 1.2 m high, and 1.2 m wide.
Vertical profiles of wind velocity and blown-sand mass flux

Wind velocity profiles and vertical sand mass flux distributions were measured using two types of sands (i.e., sand A and B) with different particle diameter compositions at five free-stream wind velocities (Table 1 and Figure 1). The mean particle diameter \(d\) and the sorting \(\sigma\) are 0.17 mm and 0.35 for sand A and 0.35 mm and 0.60 for sand B, respectively. Wind velocity was measured along the centerline of the test section entrance. During the experimental runs, the wind tunnel floor was covered with a 0.06 m thick sand bed.

Wind velocity \(u(z)\) was measured using a single hack tube connected to a digital pressure meter. A single hack tube consists of two alloy tubes (3 mm outer diameter, 2.4 mm inner diameter). The shear velocity \(u_s\) and the roughness length \(z_0\) were estimated from a semi-logarithmic regression fit to the measured velocity data obtained above the 0.03 m elevation in order to remove the effect of sand particle motion on the wind flow. Values of \(u_s\) and \(z_0\) listed in Table 1 were calculated by authors, and these values were slightly different from those obtained by NI ET AL. (2002). \(z_0\) had a clear relationship with \(u_s\), such that

\[
z_0 = c_0 \frac{u_s^2}{2g}
\]

where \(c_0\) is CHARNOCK (1955) constant (= 0.04 for experiments; RAO PACH, 1991).

Vertical sand mass flux was measured using a Liu-type passive vertical array sand trap (LIU, 1995). The trap was 0.3 m high and had 30 collection chambers. The aperture of each chamber was 0.01 m wide and 0.01 m high. The flux measured using the lowest aperture of the trap array was not accurate since traps distort the flow field near the bed. Gradients of semi-logarithmic mass flux profiles increased with both the wind velocity and the sand particle diameter; however, the gradients for sand B were relatively independent of the wind velocity. Vertical concentration profile \(C(z)\) obtained from the wind tunnel experiments by dividing \(q(z)\) by \(u_s(z)\) (equation (1)) was normalized by \(C(0)\) in order to compare it with Rouse’s concentration profile (equation (2)). In order to normalize \(C(z)\), it is critical to determine \(h\) for aeolian sand transport because in air, there is no boundary such as a water surface. Modes of sand particle motion are classified on the basis of the Rouse number \(N\) for both fluvial and aeolian sand transports (in water and air, respectively; VAN RIJN, 1993; DADE AND FRIEND, 1998; and SHAO, 2000) and VAN RIJN (1993) related \(h\) to \(N\) for fluvial transport; therefore, \(h\) was determined using the relationship between \(h\) and \(N\) for aeolian transport.

Table 2 lists the modes of sand particle motion for fluvial and aeolian transports, as suggested by VAN RIJN (1993) and SHAO (2000) under the experimental conditions for sands A and B (NI ET AL., 2002).

Vertical profile of mean blown-sand particle velocity

DONG ET AL. (2006) measured the mean sand particle velocity and concentration by particle image velocimetry (PIV) with three sand particle diameters of 0.1–0.2, 0.2–0.3, and 0.3–0.4 mm and at four free-stream wind velocities of 8.0, 10.0, 12.0, and 14.0 m s\(^{-1}\); they developed the following empirical equation of the profile of the mean particle velocity \(u_p(z)\):

\[
u_p(z) = a_4 \left( \frac{z}{Z} \right)^{b_1} U
\]

where \(a_4\) and \(b_1\) are the regression coefficients, and \(Z\) is the height of the boundary layer (=0.12 m). From the experimental results, it was found that the value of \(a_4\) was 0.90 for sand A and 0.57 for sand B, and the value of \(b_1\) was 0.5 for both sand A and B.

The value of \(u_p(z)\) calculated from equation (6) exceeded the wind velocity \(u(z)\) for \(z > 0.1\) m in cases 17A, 17B, and 17C, for all values of \(z\) in cases 17D and 17E, and for \(z > 0.27\) m in cases 35A, 35B, 35C, 35D, and 35E (Figure 1). This result is most likely true because this equation was verified only when 8 m s\(^{-1}\) < \(U\) < 14 m s\(^{-1}\) and \(z > 0.12\) m. \(u_p(z)\) seems to become close to \(u(z)\) with an increase in the wind velocity; therefore, \(u_p(z)\) was given by \(u(z)\) when \(u_p(z) > u(z)\).

Vertical profile of dimensionless concentration

The vertical concentration profile \(C(z)\) obtained from the wind tunnel experiments by dividing \(q(z)\) by \(u_p(z)\) (equation (1)) was normalized by \(C(0)\) in order to compare it with Rouse’s concentration profile (equation (2)). In order to normalize \(C(z)\), it is critical to determine \(h\) for aeolian sand transport because in air, there is no boundary such as a water surface. Modes of sand particle motion are classified on the basis of the Rouse number \(N\) for both fluvial and aeolian sand transports (in water and air, respectively; VAN RIJN, 1993; DADE AND FRIEND, 1998; and SHAO, 2000) and VAN RIJN (1993) related \(h\) to \(N\) for fluvial transport; therefore, \(h\) was determined using the relationship between \(h\) and \(N\) for aeolian transport.

Table 2 lists the modes of sand particle motion for fluvial and aeolian transports, as suggested by VAN RIJN (1993) DADE AND FRIEND (1998), and SHAO (2000). \(N\) for both the transports decreased with a mode transition from saltation to suspension; \(N\) for the modes of aeolian transport (SHAO, 2000; i.e., saltation, modified saltation, short-term suspension, and long-term suspension) roughly corresponded with that for the modes of fluvial transport (VAN RIJN, 1993; i.e., bedload (\(z < 0.1h\)), suspension (\(z < 0.5h\)), and wash load (\(z < h\)), indicating that the modes of transport could be universally
Experimental modes of NI ET AL. (2002) plotted in the figure were classified into saltation only for case 17E; however, they were classified into saltation or modified saltation for the other cases. VAN RIJN (1993) indicated that the suspension height reaches up to the water surface when $N = 1.0$; therefore, $h$ for aeolian transport was defined in terms of the height of the sand transport layer $H$ when $N = 1.0$. The relationship between $N$ and $H$ obtained from the experimental results of DONG ET AL. (2006) is shown in Figure 3. $N$ and $H$ had a negative relationship, and $H$ for $N = 1.0$ (i.e., $h$) was 0.37 m for sand A and 0.44 m for sand B, as obtained from the linear regressions.

RESULTS AND DISCUSSION

Figure 4 shows the normalized concentration profiles $C(z)/C(a)$ when $a$ was 0.05h and 100d, as measured by DONG ET AL. (2006) as well as that calculated using 100d (0.017m and 0.035 m, respectively). Inclination of $C(z)/C(a)$ tends to increase with a decrease in $N$ for both sand A and B when $a = 100d$ and for sand A when $a = 0.05h$; however, the inclination of $C(z)/C(a)$ had no clear tendency for sand B when $a = 0.05h$. It might be better to determine $a$ using a function of $d$. On the other hand, the overall $C(z)/C(a)$ for sand B was greater than that for sand A; nevertheless, $N$ for sand B was larger than that for sand A. This is most likely because there was no suspension layer for sand B with a greater $N$, or the suspension layer was excluded from $H$ when the border between the saltation and suspension layers was clearly discriminated in the experiments of NI ET AL. (2002). In addition, the value of $a$ might be large.

Comparison between $C(z)/C(a)$ obtained from the experimental results and Rouse’s profile is shown for cases 17C, 17D, and 17E in Figure 5. Experimental results of $C(z)/C(a)$ tend to become close to Rouse’s profile with a decrease in $N$ (i.e., with mode transition from saltation to suspension); however, the profile is overestimated by Rouse’s profile when $N = 1.20$, which is classified into short-term suspension (case 17E), and it corresponds with Rouse’s profile when $N = 1.62$, which is classified into modified saltation (case 17D). The profiles were also underestimated by Rouse’s profile in the case of sand B as well as cases 17A, 17B, and 17C. This should be caused by the determination methods of $u_p$, $a$, and $h$ in this study or the exclusion of the suspension layer from $H$ in the experiments of NI ET AL. (2002). The slight difference between the $N$ distributions of $D_o$ and $\Psi$ for fluvial and aeolian transports could be another cause for the underestimation of the profile. Several uncertainties still remain in the methods used to determine $u_p$, $a$, and $h$; however, it is demonstrated that the application of Rouse’s profile to aeolian transport would be possible when the mode of sand particle motion is classified into suspension.

CONCLUSIONS

In this study, a comparison was made between the normalized concentration profiles obtained from the wind tunnel experimental results by NI ET AL. (2002) and DONG ET AL. (2006) and Rouse’s suspended sediment concentration profile. The mode of sand particle motion for aeolian transport is similar to that for fluvial transport, implying that fluvial and aeolian sands have a universal suspension system. For a quantitative study, further work is required for the determination methods of the mean particle velocity and the height of suspension layer. However, in this study, Rouse’s profile became close to the experimental results with a decrease in the Rouse number, i.e., transition from saltation to suspension, and showed the best fit with the result when the

Figure 3. Relationship between $N$ and $H$ obtained from the experimental results of DONG ET AL. (2006).

Figure 4. Normalized concentration profiles obtained from the experiments (NI ET AL., 2002) when $a$ was (a) 0.05h and (b) 100d.
Rouse number was 1.62. This finding gives us a new insight into the aeolian suspended sand transport.

LITERATURE CITED


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